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AIR MINISTRY

METEOROLOGICAL RESEARCH COMMITTEE

Vorticity and Synoptic Development

by
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1. Purpose.

In the work of the Forecast Division of the Meteorological Office particular attention is given to published research or other available knowledge which offers any prospect of an improvement in the standard of forecasting. Over some years now such attention has been given to the practical application of the ideas of Sutcliffe (1947) and of new techniques, based on them, suggested by Sutcliffe and Forsdyke (1950). Experience has shown that while illuminating and valuable as a background in synoptic analysis and prognosis, nevertheless these ideas in practice sometimes confront the forecaster with puzzling discrepancies, as when, for example, pressure rises over a wide area covered by a cold trough in the upper air. It is the purpose of this paper first to illustrate difficulties, then to suggest possible reasons for them and finally to extend the discussion of the scalar vorticity equation.

2. The Sutcliffe Development Term.

Sutcliffe (loc. cit.) derived an expression for the relative divergence between pressure levels as follows:

$$f \, \mathrm{div}_p \underline{V} - f \, \mathrm{div}_p \underline{V}_0 = -V' \frac{\partial}{\partial s}(f + \zeta + \zeta_0) \qquad (1)$$

where f is the Coriolis term, $\underline{V}$ and $\underline{V}_0$ are wind vectors at two pressure levels in the atmosphere, V' is the thermal wind speed between these two levels, $\partial/\partial s$ represents differentiation along the thermal wind and ζ and ζ_0 are the vorticities at the two levels.

This expression is taken as a measure of synoptic development, following the motion of the atmosphere, and Sutcliffe suggests that the levels of 1000 mb. and 500 mb. may be used, though disclaiming any particular merit in that choice.

As a matter of interest it was thought desirable to consider the effect of varying the thicknesses. Accordingly, for 4th February, 1951, when a depression of very rare intensity affected southern Ireland the development given by equation (1) was computed for each of the thicknesses 1000-700 mb., 1000-500 mb., and 1000-300 mb. and for each of the hours 03,09, 15 and 21 G.M.T. - the method of Matthewman and Sawyer (1951) being employed. Isopleths of development were constructed, the unit being 10^{-2} hour^{-2}, and these are shown in Figures 1 to 12. The isopleths are shown by solid lines, the relevant surface isobars by broken lines, and, on the 1000-500 mb. charts, the thickness isopleths of that thickness by dotted lines. Cyclonic development is shown by negative values of the development.

/Referring

Referring first to the charts for 0300 G.M.T., we see that the depression west of Ireland was located at the northern tip of a warm tongue in the 1000-500 mb. thickness pattern, with a developing cold trough to the southwest of it. The area of maximum cyclonic development was apparently well in the cold air, with a subsidiary such maximum over the front, over southwest Ireland. On the 1000-700 mb. thickness chart there were similar centres of maximum cyclonic development in roughly similar areas; they were of about the same intensity except that the one over southwest Ireland was apparently of greater extent. On the 1000-300 mb. thickness chart the cyclonic maxima are increased relative to those on the 1000-500 mb. chart, and, while that over southwest Ireland is again broadly in the same location as those on the other two charts, the one on the Atlantic is appreciably nearer the centre of the depression than the corresponding one at 500 mb.

By 0900 G.M.T. the centres of the depression had moved in over western Ireland, the lowest pressure having fallen by two millibars. On the 1000-700 thickness chart the area of maximum cyclonic development is near the Isle of Man and again in proximity to the major front of the depression. The subsidiary maximum on the Atlantic has become much less intense but more extensive. On the 1000-500 mb. chart the area of maximum cyclonic development is over northwest Ireland, almost over the main centre of the depression, and its value is similar to that shown on the Atlantic on the earlier corresponding chart. On the 1000-300 mb. chart the area of maximum cyclonic development is appreciably further west than the corresponding area on the 1000-500 mb. chart, its value being similar to that for the same thickness six hours earlier.

At 1500 G.M.T. the depression was centred over the heart of Ireland, its central pressure being lower than 944 mb. Apparently the tendency for a double centre was dying out. The thickness lines on the 1000-500 mb. chart show a pool of cold air south of the centre of the low. The 1000-700 mb. development chart shows what is apparently a decaying cyclonic maximum over northeast Ireland with an anticyclonic maximum off southwest Ireland. The corresponding chart for 1000-500 mb. shows an area of cyclonic development from north of Stornoway southwards over Ireland and another to the southwest of Ireland. The intensity of this development is much less than the maximum of six hours earlier. On the 1000-300 mb. development chart the area of maximum cyclonic development is off northwest Ireland and, again, its intensity is reduced relative to the corresponding maximum of six hours earlier. As on the thickness chart for 1000-500 mb., there is a subsidiary maximum of cyclonic development off southwest Ireland, though further west and north.

By 2100 G.M.T. the depression was centred over eastern Ireland and the Irish Sea and the area of maximum cyclonic development on the 1000-700 mb. chart was just ahead of it over northern England. On both the charts for 1000-500 mb. and 1000-300 mb., however, the area of maximum cyclonic development is shown to the northwest of the depression, apparently in the relatively cold air. This tendency for the area of maximum cyclonic development to retrogress into the cold air as the thickness dealt with is increased is shown on all the charts except those for 0300 G.M.T. and even on these the conclusion must be somewhat doubtful. We see then that the region of maximum cyclonic development varies with the thickness of the atmosphere considered.

Restrictive Conditions.

McVittie (1952) has dealt at length with the restrictive conditions which apply in the derivation of the development term but it may be of interest to proceed from Sutcliffe's own equations to see where we arrive.

/To

To obtain equation (1), Sutcliffe (loc. cit.), on the assumption that the sum of

$$-\xi \operatorname{div}_p \underline{V} - \frac{\partial v}{\partial p}\left(\frac{\partial}{\partial x}\frac{dp}{dt}\right)_p + \frac{\partial u}{\partial p}\left(\frac{\partial}{\partial y}\frac{dp}{dt}\right)_p$$

is negligible derives the approximate scalar vorticity equation:

$$\frac{d}{dt}(\xi + \ell) = -\ell \operatorname{div}_p \underline{V} \qquad \text{(2)}$$

Eady (1952) says: "This is certainly not true in the vicinity of a "jet" and the relative error may also be large near intense depressions or anticyclones. In general the errors are larger at high than at low levels ____". It may be that these words of Eady's provide the explanation for the apparent inconsistencies found in this study.

However, proceeding from the approximate equation (2) we find:

$$\frac{d\xi}{dt} - \frac{d\xi_0}{dt} = -\ell\left(\operatorname{div}_p \underline{V} - \operatorname{div}_p \underline{V}_0\right) \qquad \text{(3)}$$

but

$$\frac{d\xi}{dt} - \frac{d\xi_0}{dt} \equiv \frac{\partial \xi'}{\partial t} + \underline{V}.\nabla(\xi + \ell) - \underline{V}_0.\nabla(\xi_0 + \ell)$$

$$= \frac{\partial \xi'}{\partial t} + (\underline{V}_0 + \underline{V}').\nabla(\xi_0 + \xi' + \ell) - \underline{V}_0.\nabla(\xi_0 + \ell)$$

$$= \frac{\partial \xi'}{\partial t} + \underline{V}_0.\nabla\xi' + \underline{V}'.\nabla\xi_0 + \underline{V}'.\nabla(2\xi_0 + \xi' + \ell)$$

From equations (1) and (3) we see that the left-hand side of this equation and the last term on the right-hand side are equal, so we are left with:

$$-\frac{\partial \xi'}{\partial t} = \underline{V}_0.\nabla\xi' - \underline{V}'.\nabla\xi_0$$

or, since the direction of shear is constant with height, following Sutcliffe:

$$-\frac{\partial \xi'}{\partial t} = V_0\frac{\partial \xi'}{\partial s} - V'\frac{\partial \xi_0}{\partial s} \qquad \text{(4)}$$

This then is a condition of the expression for development in equation (1), namely, that local change of thermal vorticity is due to the difference in advection of thermal vorticity by the surface wind and of surface vorticity by the thermal wind. An alternative which may be implicit in Sutcliffe's analysis is to regard this equation as a residual, negligible in comparison with the terms he takes into account but on the other hand the importance of the term $V'\frac{\partial \xi_0}{\partial s}$ - the "thermal steering effect" - is stressed.

/The

The Vertical Variation terms.

Sutcliffe (1947) derives the vorticity equation in the form:

$$\frac{d}{dt}(\zeta+\ell) = -(\zeta+\ell)\,\mathrm{div}_p \underline{V} - \frac{\partial v}{\partial p}\left(\frac{\partial}{\partial x}\frac{dp}{dt}\right)_p + \frac{\partial u}{\partial p}\left(\frac{\partial}{\partial y}\frac{dp}{dt}\right)_p$$

whence

$$\frac{d}{dt}(\zeta+\ell) = -(\zeta+\ell)\,\mathrm{div}_p \underline{V} - \nabla_p \frac{dp}{dt} \times \frac{\partial \underline{V}}{\partial p} \qquad (5)$$

Sutcliffe (1947) argues that the sum: $\left(\zeta\,\mathrm{div}_p \underline{V} + \nabla_p \frac{dp}{dt} \times \frac{\partial \underline{V}}{\partial p}\right)$ is negligible but Sherman (1952) has shewn that terms equivalent to $\nabla_p \frac{dp}{dt} \times \frac{\partial \underline{V}}{\partial p}$ are of the same order of magnitude as the sum of the advective terms: $\left[\underline{V}.\nabla_p(\zeta+\ell) + \frac{dp}{dt}\frac{\partial}{\partial p}(\zeta+\ell)\right]$ and the term of the form $(\zeta+\ell)\,\mathrm{div}_p \underline{V}$. Reed (1951) made some computations to show that both the terms on the right-hand side of equation (5), or term of that type, were of equal importance. McVittie (1952) also shews that the magnitude of these two terms are comparable and that both should be retained in the scalar vorticity equation.

The attempt was made to evaluate the terms on the right-hand side of equation (5) on two occasions of special interest and the results are given in Table I. Even if the computations proved unreliable as to the exact values of the terms, it was hoped that at least they would give some guide as to their relative order of magnitude. Divergence and vorticity were computed by Bellamy's (1949) "objective" method, the values in Table I being given in units of hour^{-2}.

Values found for $\nabla_p \frac{dp}{dt} \times \frac{\partial \underline{V}}{\partial p}$ compare well with those found by Reed (1951) for his equivalent term, namely, 10^{-9} to 10^{-8} sec^{-2}. In the table it may also be seen that the sum of $\zeta\,\mathrm{div}_p \underline{V}$ and $\nabla_p \frac{dp}{dt} \times \frac{\partial \underline{V}}{\partial p}$ is not usually zero and that $-\ell\,\mathrm{div}_p \underline{V}$ is not always the highest absolute magnitude. Eady's (1952) point that errors due to the approximation would be larger at higher levels is also confirmed.

When the values of the individual time rate of change of the vorticity at 1000 mb. relative to that at 500 mb. are plotted, the charts obtained are as shewn in Figures 13 and 14. By comparing Figure 13 with Figures 1 and 4 it may be seen that the area of increasing cyclonic vorticity over southern Ireland at 0300 G.M.T. was followed at 0900 G.M.T. by the development of a small secondary in that area, while the area of increasing vorticity over eastern England was associated with the major front of the depression. In Figure 14 the area of increasing cyclonic vorticity off southwest Ireland was followed by the advance into that area of a depression from the Atlantic with its associated front advancing into southwest England. To the extent illustrated by these two examples the computations may be taken as giving real results.

/The

The term Curl $\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right)$

Returning to equation (5), we get, by transferring the term $\frac{dp}{dt}\frac{\partial}{\partial p}(\zeta+\ell)$ from the left to the right-hand side:

$$\left[\frac{d}{dt}(\zeta+\ell)\right]_p = -(\zeta+\ell)\,\mathrm{div}_p \underline{V} - \mathrm{curl}_p\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right) \qquad (6)$$

This reduces to:

$$\left[\frac{d}{dt}(\zeta+\ell)\right]_p = -(\zeta+\ell)\,\mathrm{div}_p \underline{V} \qquad (7)$$

if:

(a) the motion is barotropic, though in this case the term on the right-hand side is small.

(b) the thermal wind is zero, i.e. if the motion is "solid". This is equivalent to (a).

(c) the thermodynamical development represented by $\frac{dp}{dt}$ is zero, i.e. if the vertical motion is zero.

(d) if $\mathrm{Curl}_p\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right) \equiv 0$ neither $\frac{\partial \underline{V}}{\partial p}$ nor $\frac{dp}{dt}$ being zero. In this event $\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}$ must be the gradient of a scalar point function, its direction being determined by that of the thermal wind.. Then also the vorticity equation is similar to that of "solid" motion or of motion without dynamical development. We can say, defining $P \equiv -\mathrm{curl}\left(\frac{\partial \underline{V}}{\partial p}\frac{dp}{dt}\right)$ —

(1) On an isobaric surface where there is no divergence, the sign and magnitude of P will determine those of the individual time rate of change of absolute vorticity.

(2) Generally, the individual time rate of change of absolute vorticity will be modified by P, which, when positive will contribute cyclonic vorticity to the change, and, when negative will contribute anticyclonic vorticity to the change.

The magnitude of the contribution of P to the time rate of change of vorticity on any particular occasion cannot readily be estimated since it is a function of $\frac{dp}{dt}$, perhaps the most important unknown in the equations of motion. However, we may arrive at a qualitative estimate of the effect of P if we consider it in its other form:

$$P = -\mathrm{curl}_p\left(\frac{\partial \underline{V}}{\partial z}w\right) = \frac{\partial \underline{V}}{\partial z} \times \nabla_p w - w\,\mathrm{curl}\frac{\partial \underline{V}}{\partial z} \qquad (8)$$

where w is vertical motion relative to the pressure surface and $\frac{\partial \underline{V}}{\partial z}$ is the thermal wind. For reference we may call the first of the two terms on the right-hand side of the identity (8) P_1 and the second, together with its minus sign P_2 so that $P = P_1 + P_2$.

/Positive

Positive and negative values of w will be shewn respectively by areas of continuous rainfall and by areas of subsidence, the former normally being associated at levels near the surface with cyclonic vorticity and the latter with anticyclonic vorticity. A qualitative estimate of P may be made by reference to such areas. The positive direction of the gradient of w will be shown by the perpendiculars to the boundaries of these areas, from the area of no-rain towards the rain and from the area of subsidence towards that of no-subsidence. The thermal wind vector is taken as positive in the direction of the thermal wind. Then a selection of possible cases may be considered:-

(1) w positive - i.e. cyclonic vorticity near the surface.

(a) P_1 positive - i.e. air rising at an increasing rate from right to left of the thermal wind. P_2 will be positive if the vorticity of the thermal wind is anticyclonic, and the contribution of P to the change in vorticity will be positive (or cyclonic). It is possible to envisage such a series of conditions near a warm front on the flank of a thermal ridge and they may have relevance in the formation of warm front "breakaway" depressions, though this requires to be verified.

(b) P_1 negative - i.e. air rising at an increasing rate from left to right of the thermal wind. P_2 will be negative if the vorticity of the thermal wind is cyclonic and the contribution of P to the change in vorticity will be negative (or anticyclonic). Such a series of conditions could occur near a cold front on the flank of a thermal trough - with a resulting ridge of high pressure developing behind it.

(2) w negative - i.e. anticyclonic vorticity near the surface.

(a) P_1 positive - i.e. air subsiding at an increasing rate from left to right of the thermal wind. P_2 will be positive if the vorticity of the thermal wind is cyclonic and the contribution of P to the change in vorticity will be cyclonic. Such a series of conditions could occur on the periphery of an anticyclone underlying an upper thermal ridge, with waves passing round the anticyclone.

(b) P_1 negative - i.e. air subsiding at an increasing rate from right to left of the thermal wind. P_2 will be negative if the vorticity of the thermal wind is anticyclonic and the contribution of P to the change in vorticity will be anticyclonic. Such a series of conditions could occur with the extension of an anticyclone or ridge of high pressure.

(3) Constant thermal wind, - i.e. $\mathrm{Curl}_p \frac{\partial \underline{V}}{\partial z} = 0$.

In this case the contribution of P to the change in vorticity will be cyclonic with subsiding air on the warm side of the thermal wind, as on the south side of a west - east airstream, or with rising air on the north side of such a stream. With rising air to the south or subsiding air to the north then the contribution of P would be anticyclonic.

(4) $\mathrm{div}_p \underline{V} = 0$ - i.e. w constant for the pressure thickness throughout which the thermal wind is taken. As indicated above, cyclonic or anticyclonic vorticity would develop in this case according as the sign of P is positive or negative. If w were constant also over an area then, with rising air, existing cyclonic vorticity would be increased in a region of anticyclonic thermal vorticity and diminished in a region of cyclonic thermal vorticity.

/Conclusion

Conclusion.

It is shown that seeming discrepancies in the application of Sutcliffe's (1947) ideas of development may be due to the employment of an over-simplified scalar vorticity equation and that, in particular, neglect of the vertical variation term may not be justified. This confirms the view of McVittie (1952) who appears sceptical in regard to its usual neglect. A form of this term is derived with no restrictive conditions on the motion of the atmosphere apart from those implicit in the usual two-dimensional frictionless equations of motion: $-\mathrm{Curl}_p\left(\frac{\partial V}{\partial p}\frac{dp}{dt}\right)$. Defined as P, for convenient reference, it is shewn that its contribution to the time rate of change of vorticity may be profitably considered in forecasting.

A few examples have been given of the possible influence of P and a systematic study of that influence would appear to be justified. In particular the behaviour of depressions and anticyclones relative to the jet stream could perhaps be clarified and also the behaviour of depressions in cold pools and of anticyclones in warm ridges.

As for the day to day use in forecasting of the term P, provided areas of rising and subsiding air could be delineated on the "thickness" charts, qualitative assessments could be made probably with an improvement on present forecasting results.

It may also be that P could be assessed for mean charts over extended periods, with possible gain to the forecaster concerned with such periods.

Acknowledgments.

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TABLE I

4 FEB 1951 0300 GMT

950 MB.

	55.5N 16.6W	57.3N 12.4W	55.7N 08.0W	53.3N 07.5W	56.5N 05.7W	54.9N 04.4W	52.8N 05.0W	54.2N 02.0W	52.5N 01.5W	52.6N 12.0W	51.8N 03.5W
$-\xi \operatorname{div} \underline{V}$	-0.048	+0.009	-0.014	-0.029	+0.003	-0.006	-0.022	-0.026	0	+0.028	+0.021
$-\ell \operatorname{div} \underline{V}$	-0.086	-0.044	+0.039	+0.071	+0.009	-0.052	-0.05	+0.084	+0.004	+0.029	+0.066
$-\overline{\xi+\ell} \operatorname{div} \underline{V}$	-0.134	-0.035	+0.025	+0.042	+0.012	-0.058	-0.072	+0.058	+0.004	+0.057	+0.087
$-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$	+0.006	+0.008	+0.01	+0.002	+0.01	-0.015	-0.015	-0.01	+0.015	-0.01	-0.003
$\frac{d}{dt}(\xi+\ell)$	-0.128	-0.027	+0.035	+0.044	0	-0.073	-0.087	+0.048	+0.019	+0.047	+0.084

500 MB.

$-\xi \operatorname{div} \underline{V}$	-0.008	-0.02	-0.011	-0.011	+0.01	-0.003	-0.031	0	+0.003	+0.067	-0.005
$-\ell \operatorname{div} \underline{V}$	-0.013	-0.057	+0.069	+0.025	+0.013	-0.009	-0.042	0	-0.021	+0.098	-0.008
$-\overline{\xi+\ell} \operatorname{div} \underline{V}$	-0.021	-0.077	+0.058	+0.014	+0.023	-0.012	-0.073	0	-0.018	-0.165	-0.013
$-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$	+0.07	+0.05	+0.03	-0.02	+0.08	-0.05	0	-0.03	+0.006	+0.02	+0.08
$\frac{d}{dt}(\xi+\ell)$	+0.051	-0.027	+0.088	-0.006	+0.103	-0.062	-0.073	0	-0.012	+0.145	+0.067
$\left(\frac{d\xi}{dt}\right)_{950} - \left(\frac{d\xi}{dt}\right)_{500}$	-0.179	0	-0.053	+0.05	-0.103	-0.011	-0.014	+0.048	+0.031	-0.098	+0.017

3 OCT 1952 1500 GMT

950 MB.

	57.3N 13.6W	59.8N 07.7W	55.3N 10.9W	56.5N 05.5W	58.3N 03.5W	53.2N 11.8W	53.4N 06.5W	54.8N 04.3W	56.4N 01.0W	52.0N 06.0W	54.3N 01.6W	51.8N 03.5W	52.5N 01.0W
$-\xi \operatorname{div} \underline{V}$	+.002	-.007	-.006	-.003	-.001	+.001	+.001	-.002	+.013	0	-.003	-.004	-.001
$-\ell \operatorname{div} \underline{V}$	-.022	-.077	+.022	+.026	-.044	-.004	-.008	+.017	-.039	+.012	+.008	+.008	+.008
$-\overline{\xi+\ell} \operatorname{div} \underline{V}$	-.02	-.084	+.016	+.023	-.045	-.003	-.007	+.015	-.026	+.012	+.005	+.004	+.007
$-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$	-.005	-.005	+.0009	-.002	-.005	+.003	-.0008	+.0003	+.0004	-.0018	-.0001	+.0011	-.0006
$\frac{d}{dt}(\xi+\ell)$	-.025	-.089	+.0169	+.021	-.05	0	-.0078	+.0153	-.0256	+.0102	+.0049	+.0051	+.0064

500 MB.

$-\xi \operatorname{div} \underline{V}$	+.003	+.031	-.003	+.01	-.001	0	+.001	0	-.001	-.002	+.007	+.006	-.005
$-\ell \operatorname{div} \underline{V}$	-.057	+.063	+.009	-.026	-.031	0	-.004	0	-.004	+.12	+.017	-.033	-.053
$-\overline{\xi+\ell} \operatorname{div} \underline{V}$	-.054	+.094	+.006	-.016	-.032	0	-.003	0	-.005	+.01	+.024	-.027	-.058
$-\nabla_p \frac{dp}{dt} \times \frac{\partial V}{\partial p}$	+.03	-.03	-.02	+.01	0	-.05	+.02	+.004	-.03	-.03	+.001	-.04	+.01
$\frac{d}{dt}(\xi+\ell)$	-.024	+.064	-.014	-.006	-.032	-.05	+.017	+.004	-.035	-.02	+.025	-.067	-.048
$\left(\frac{d\xi}{dt}\right)_{950} - \left(\frac{d\xi}{dt}\right)_{500}$	-.001	-.153	+.031	+.027	-.018	+.05	-.025	+.011	+.009	+.03	-.02	+.072	+.054

FIG. 1. FEB 4 1951
0300 GMT 1000-700 MB

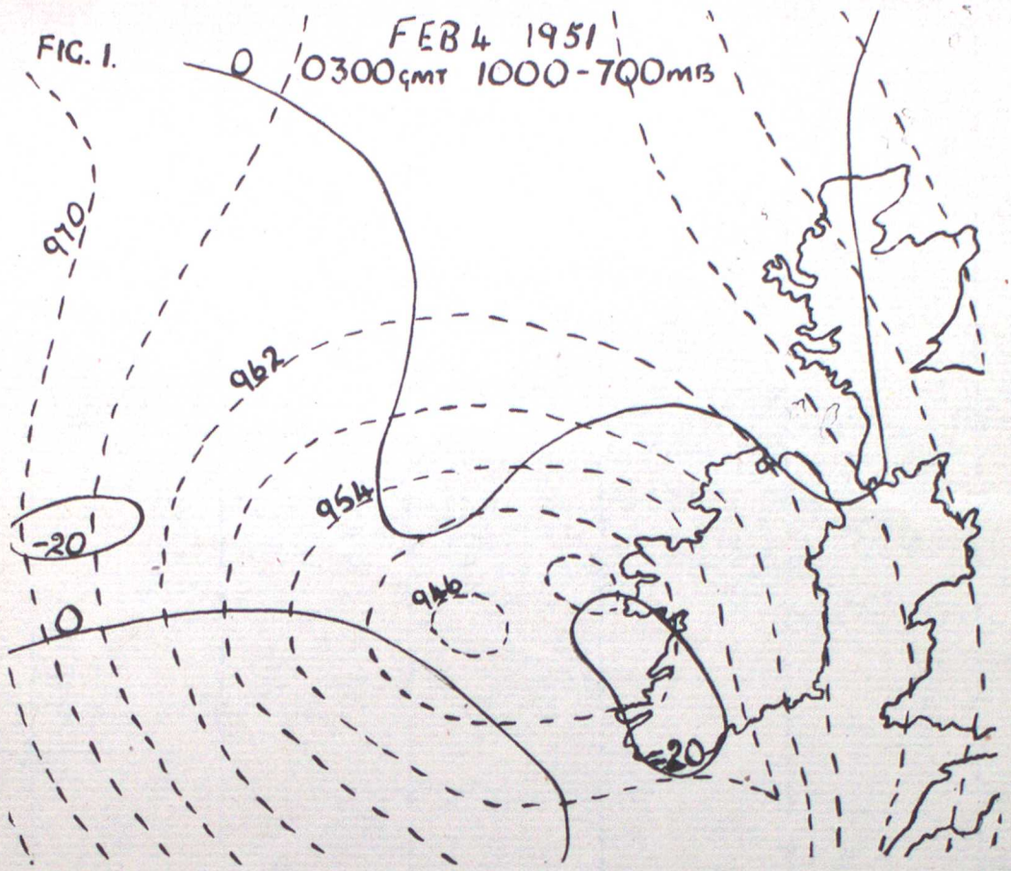


FIG. 2 FEB. 4 1951
0300 GMT 1000-500 MB

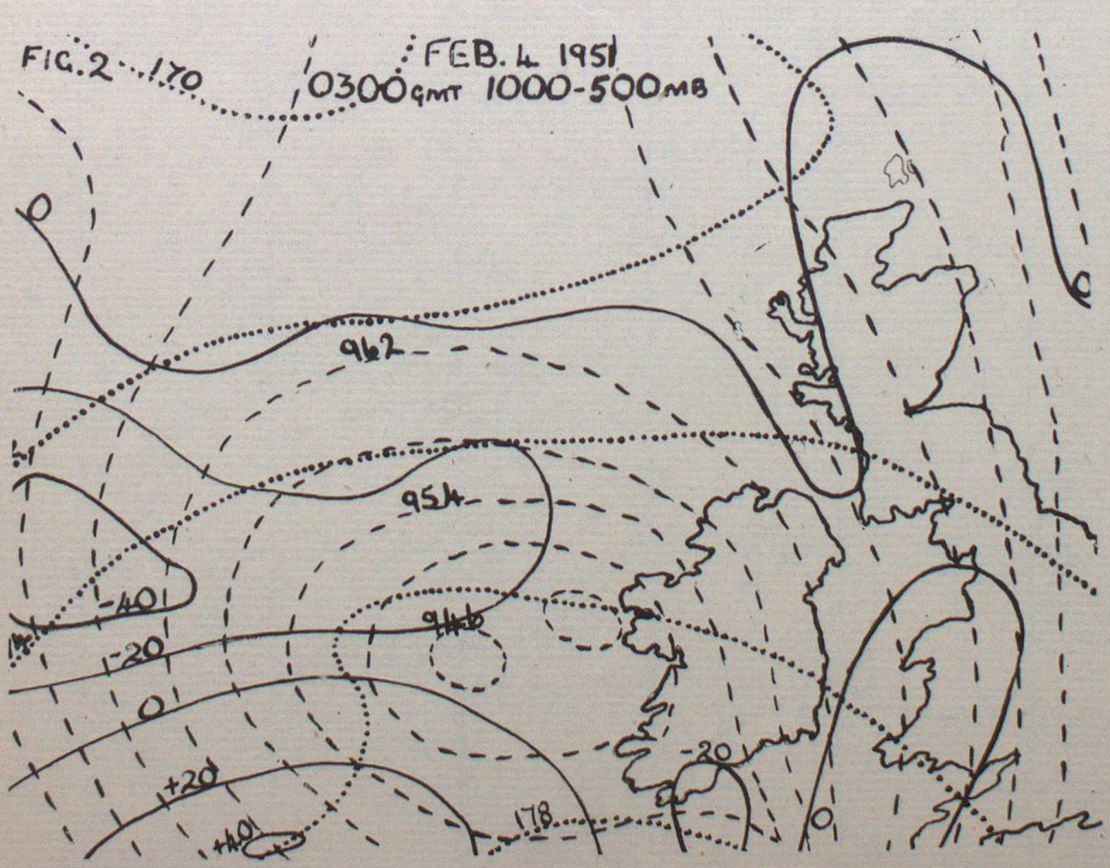


FIG. 3 FEB 4 1951
0300 GMT 1000-300 MB

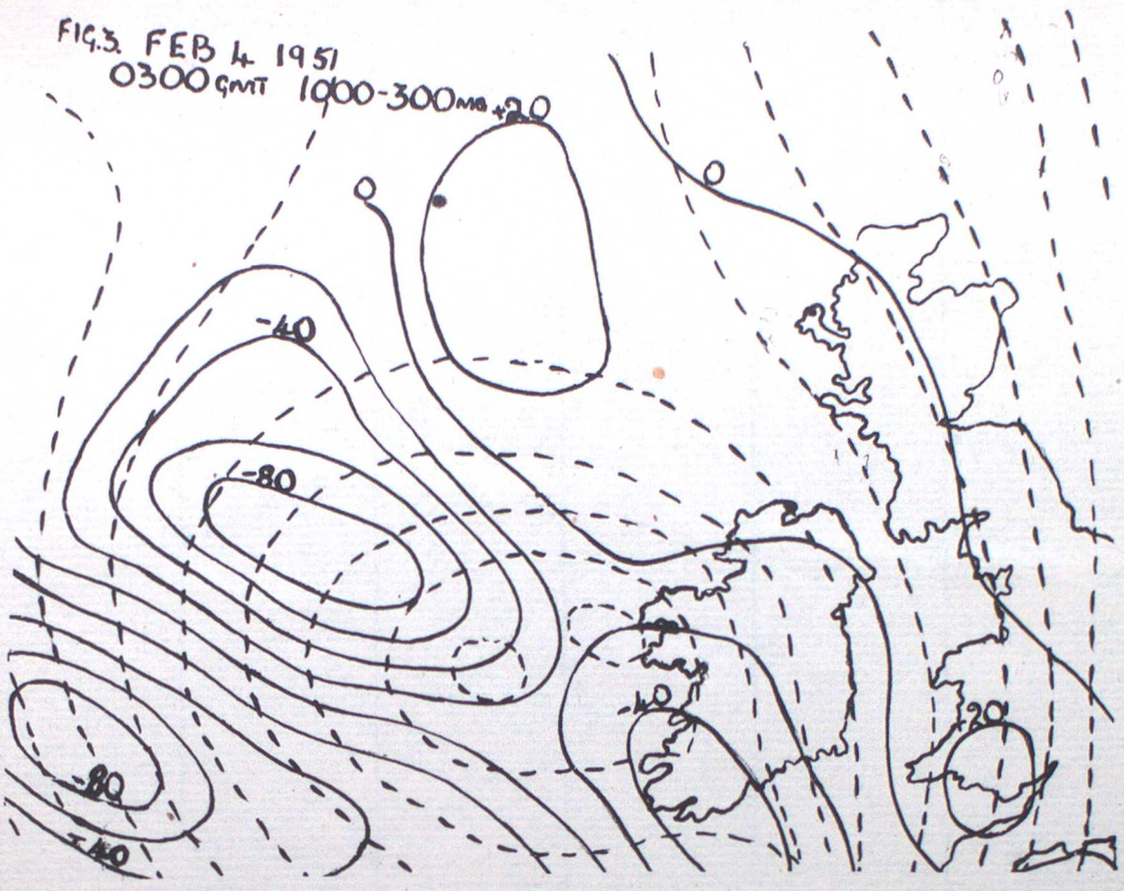


FIG. 4 FEB 4 1951
0900 GMT. 1000-700 MB

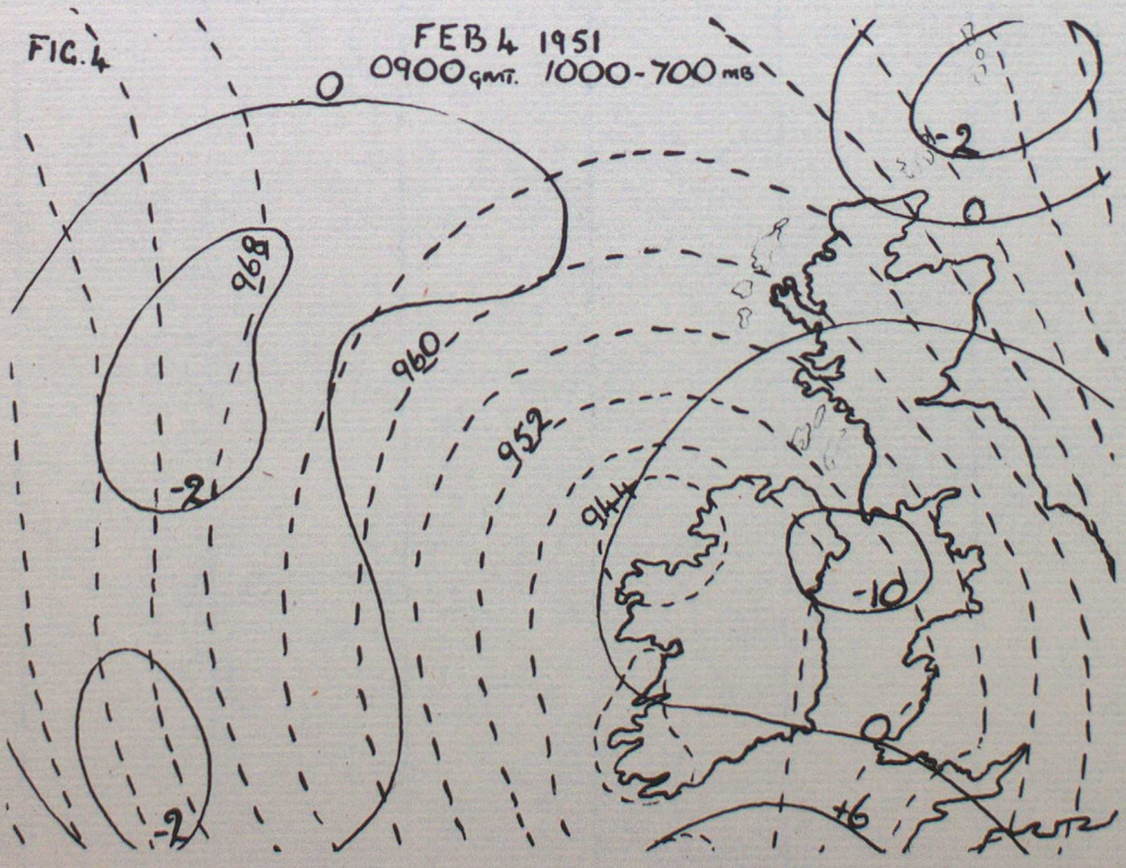


FIG. 5 FEB. 4 1951
0900 G.M.T 1000-500 MB

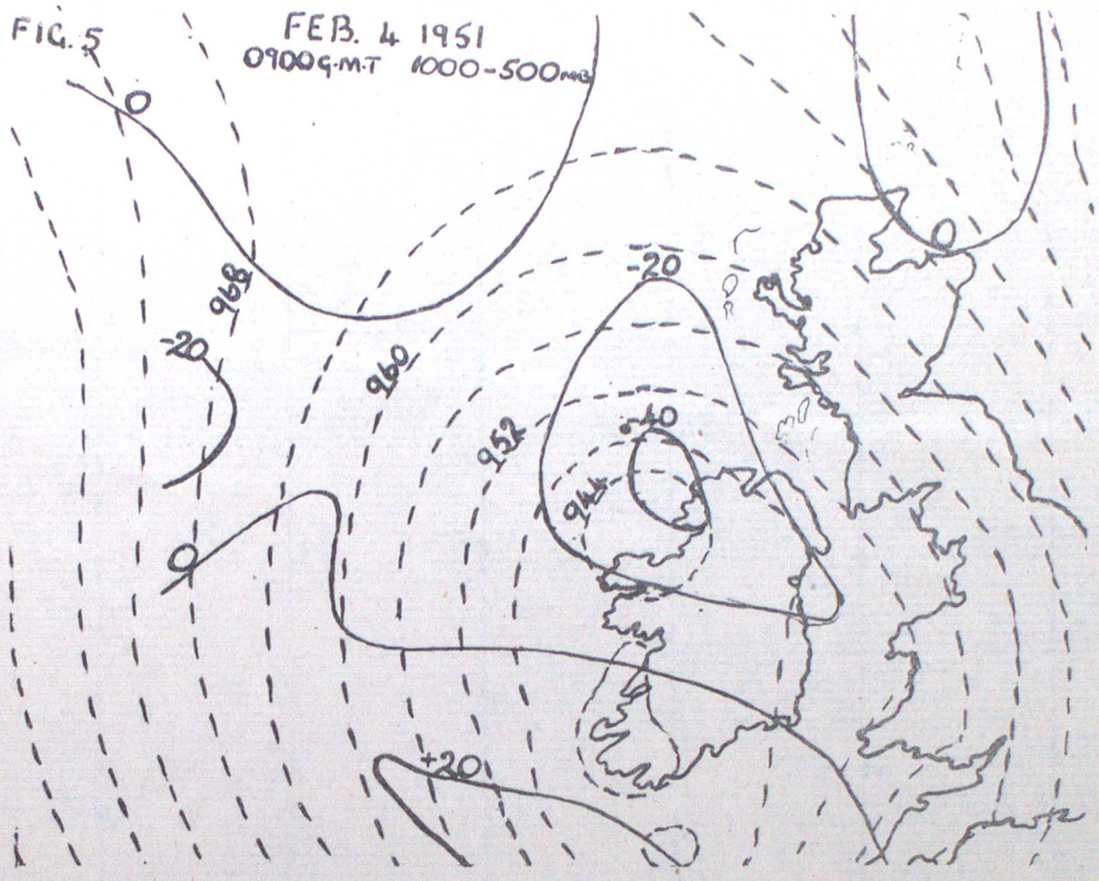
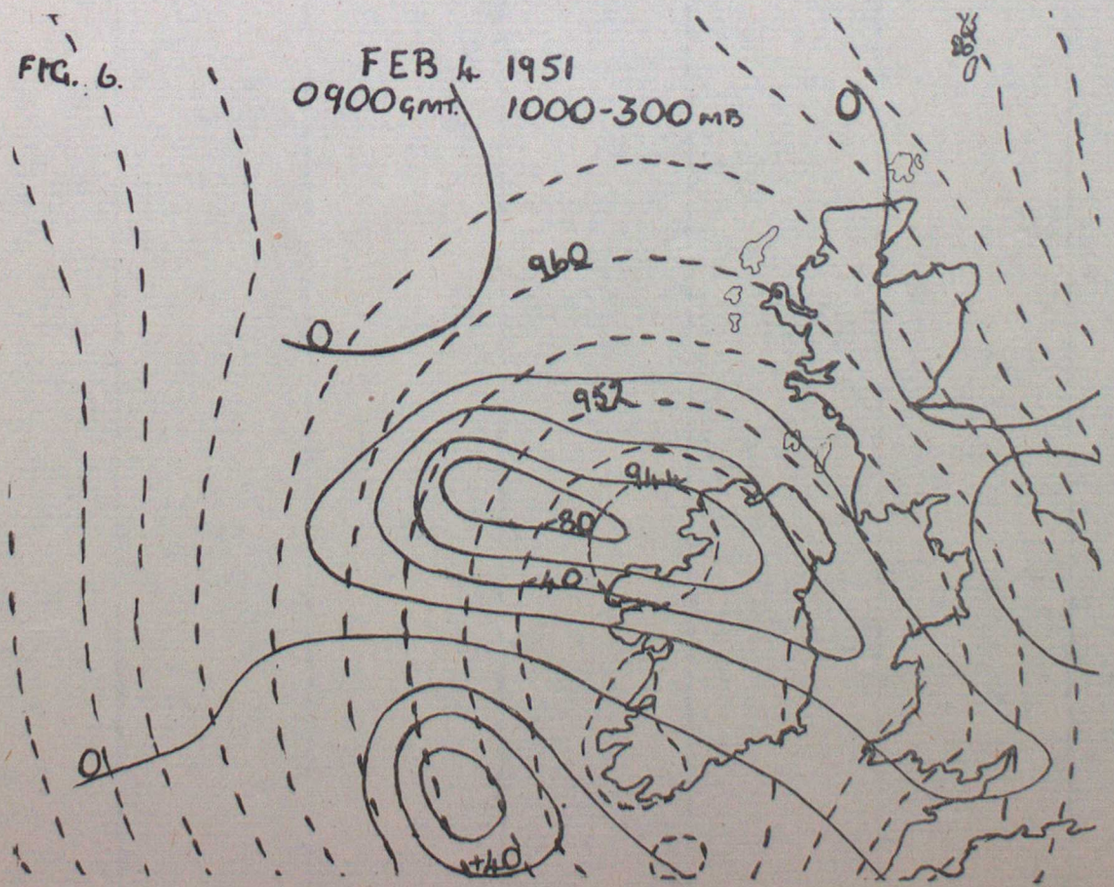


FIG. 6. FEB 4 1951
0900 GMT. 1000-300 MB



M.R.P. 829

FIG. 7. 1500 GMT FEB 4 1951 1000-700 MB

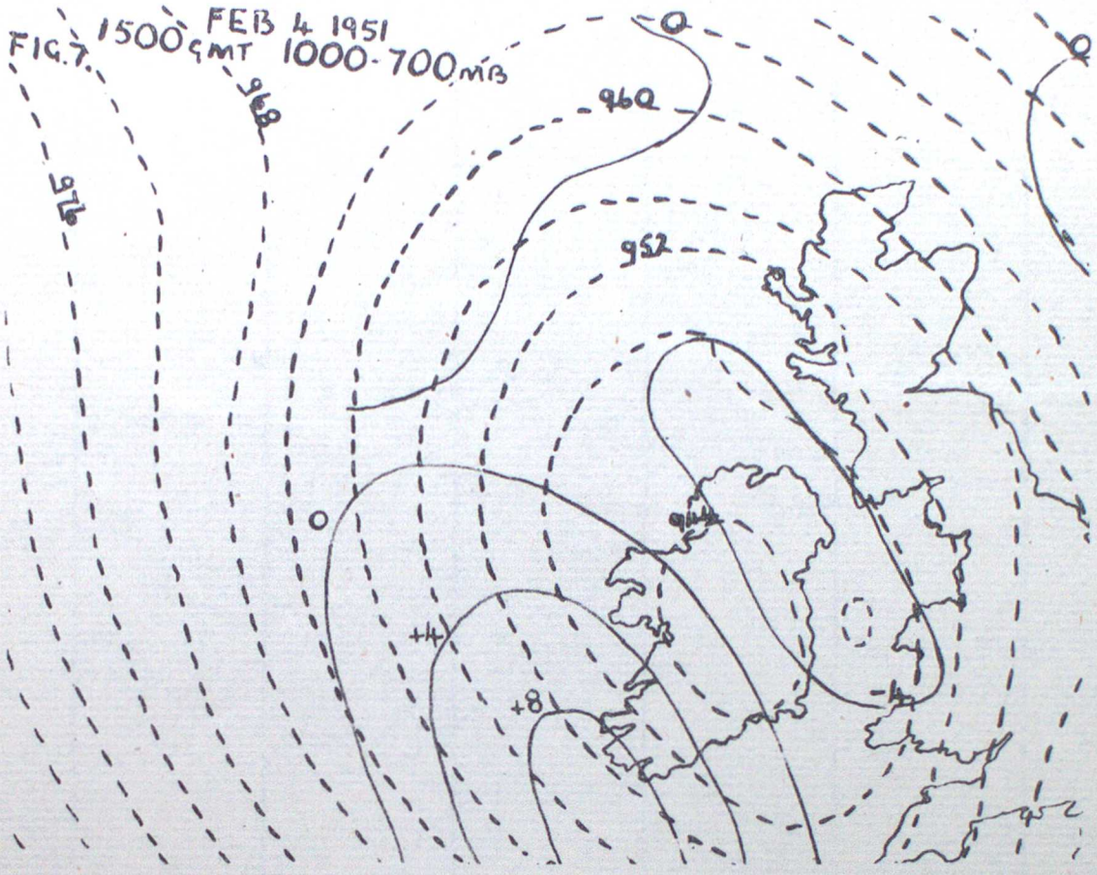


FIG. 8 1500 GMT FEB 4 1951 1000-500 MB

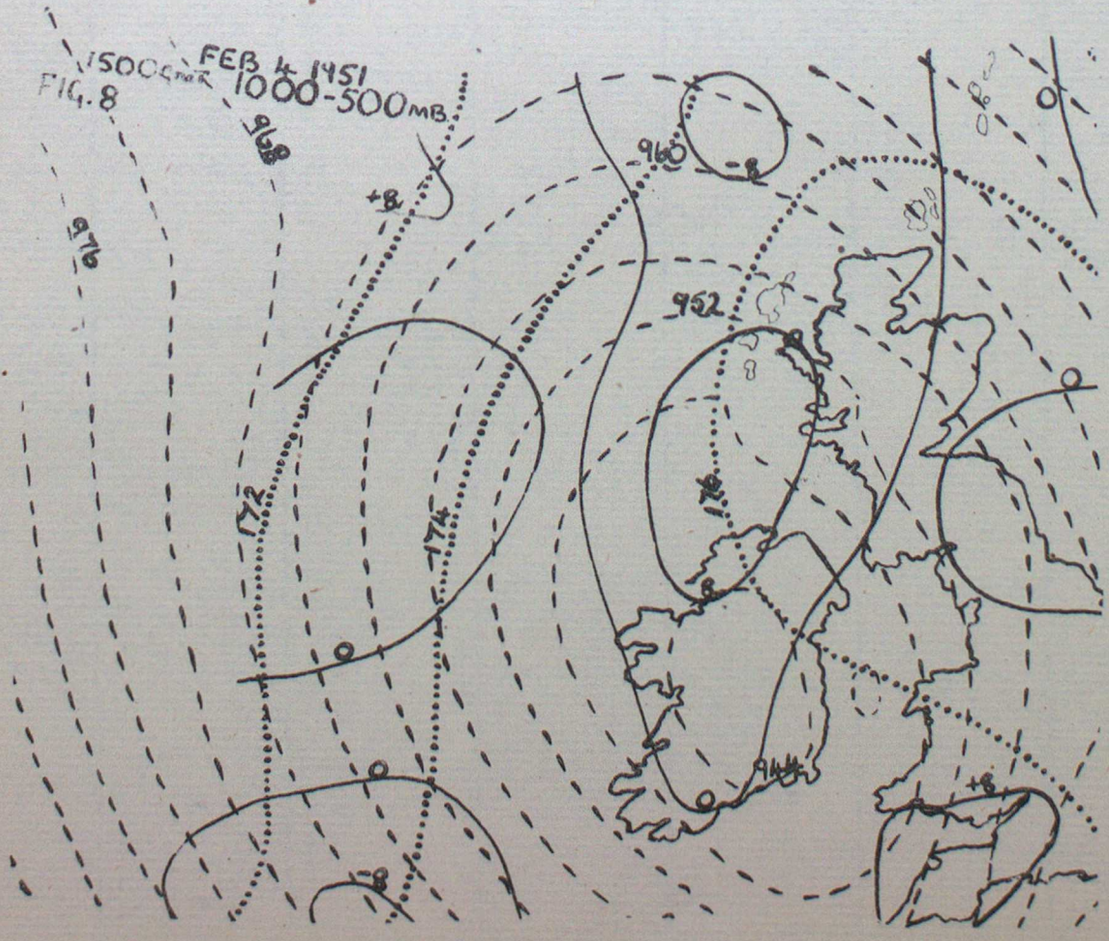


FIG. 9. FEB. 4. 1951 1500 GMT 1000-300 MB

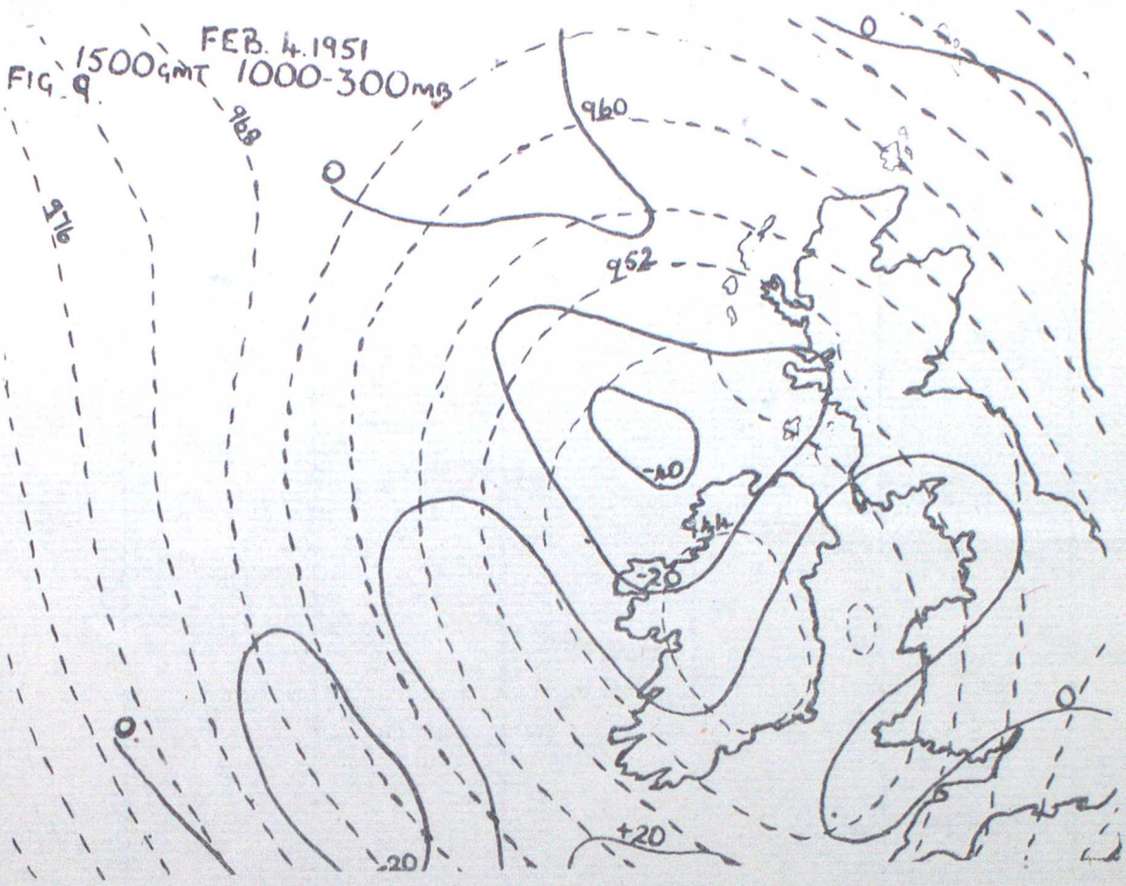


FIG. 10. FEB 4 1951 2100 GMT 1000-700 MB

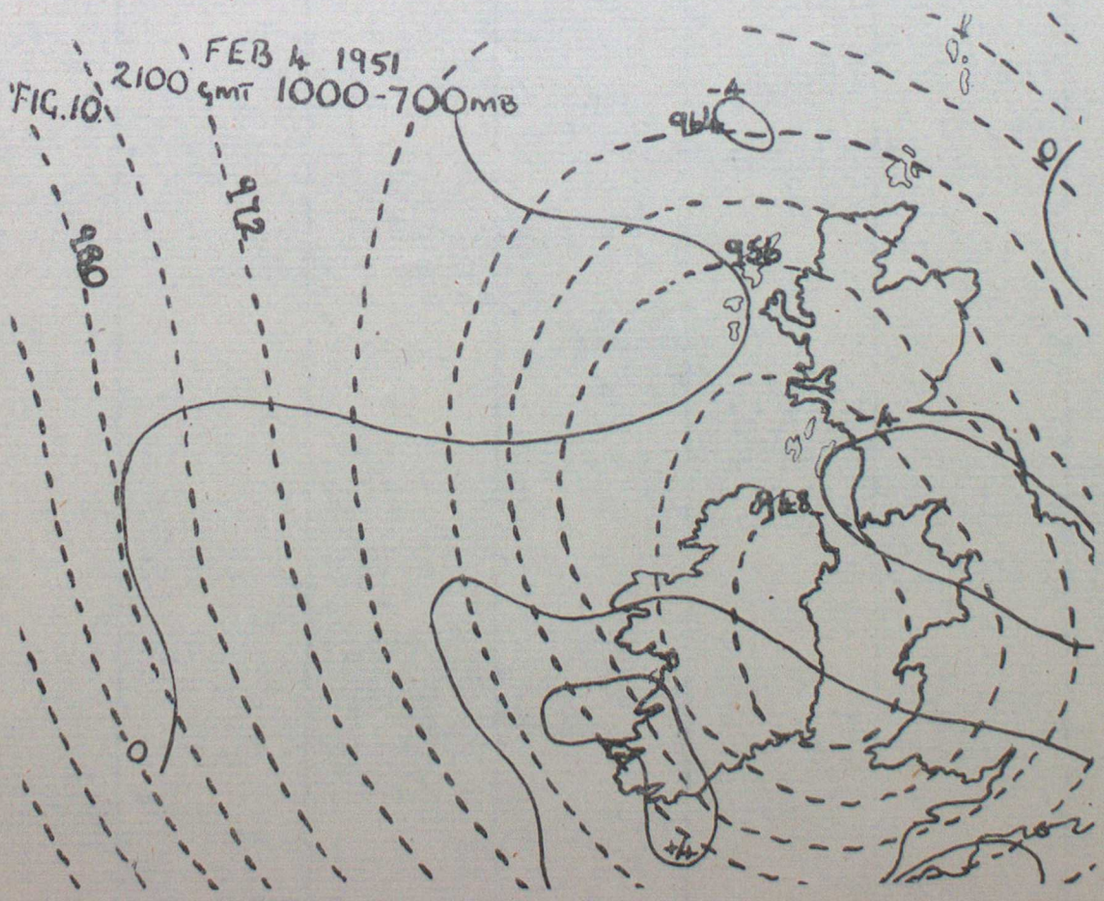


FIG. 11. FEB 4 1951 2100 GMT 1000-500 MB

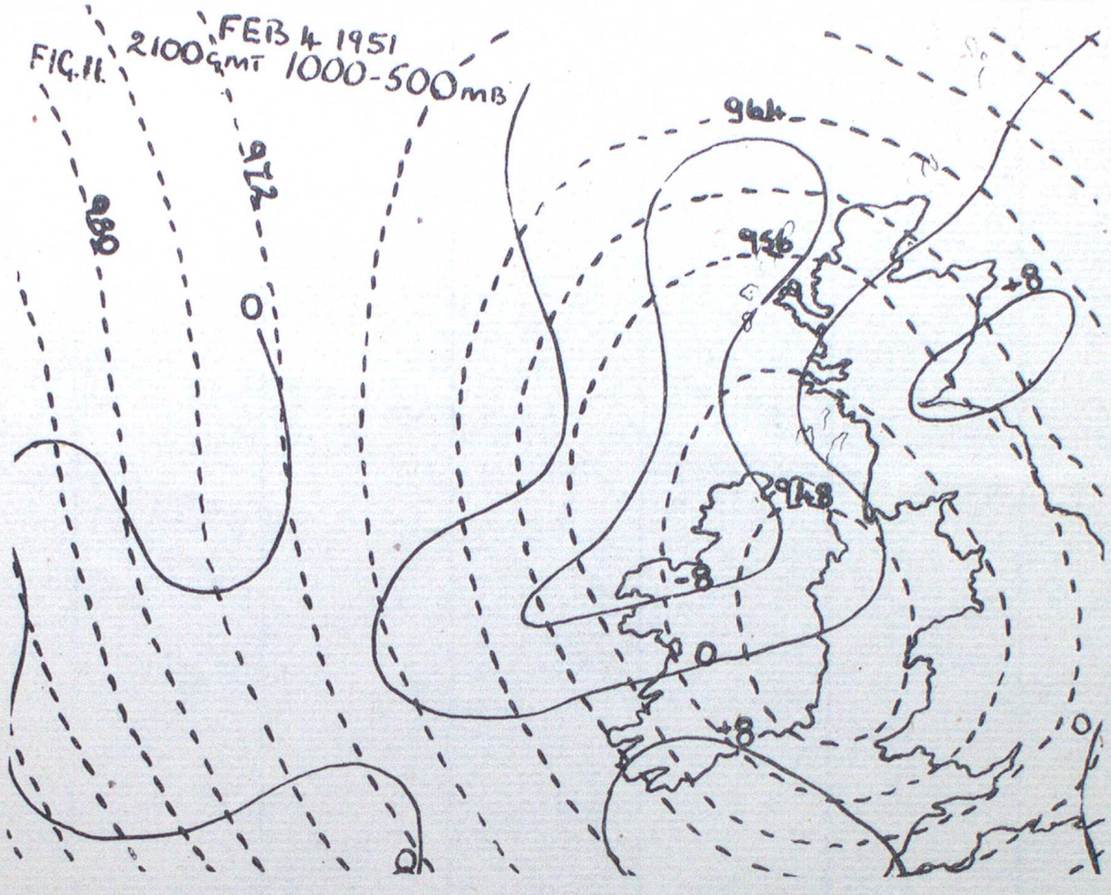
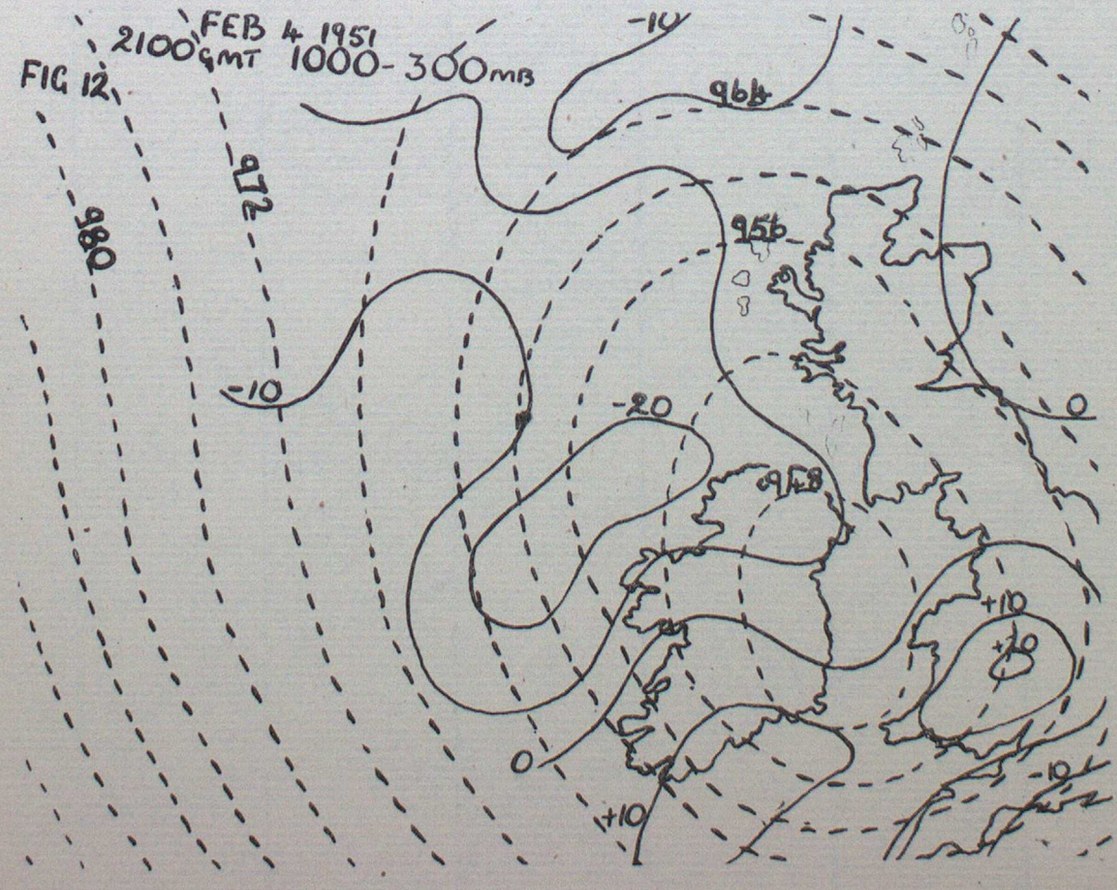


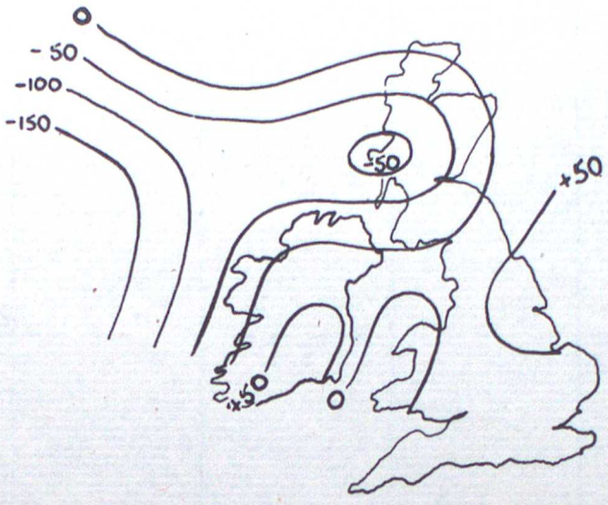
FIG. 12. FEB 4 1951 2100 GMT 1000-300 MB



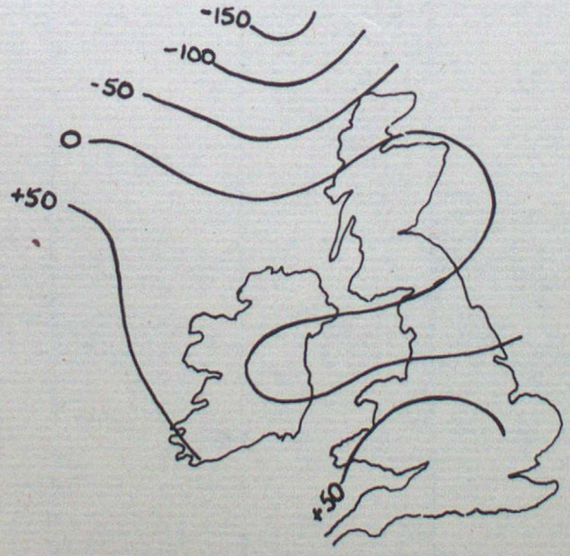
FEB 4 1951 0300 GMT

FIG 13.

$\frac{d(f+\zeta)}{dt}$ in units 10^{-3} hr^{-2}



OCT 3 1952 1500 GMT

FIG 14

$\frac{d(f+\zeta)}{dt}$ in units 10^{-3} hr^{-2}